Marco.Zavatarelli@unibo.it

Thermohaline circulation
Part 1

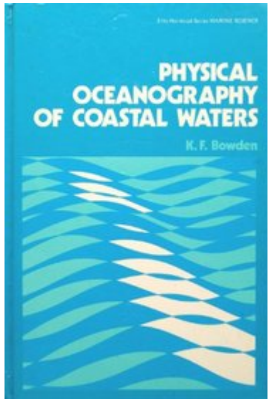
Main references

The Sea: Vol 10.
The global coastal Ocean:
Processes and methods

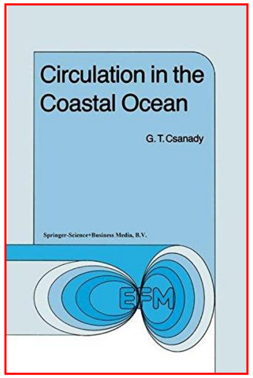
Chapter 2. BUOYANCY EFFECTS IN COASTAL AND SHELF SEAS

A. EDWARD HILL

School of Ocean Sciences
University of Wales, Bangor



K. F. Bowden.
Physical Oceanography of coastal waters:
Chapter 6: Density currents and salinity distribution.
Sections 6.1, 6.3
Chapter 7: Temperature distribution and the Seasonal thermocline
Sections 7.1, 7.2 7.3



G.T Csanady: Circulation in the coastal ocean.
Chapter 7. Thermohaline circulation
Sections 7.0, 7.1

Some definitions

- *Buoyancy (reduced gravity)*

$$b = -g\frac{\rho - \rho_0}{\rho_0} = -g\varepsilon = g'$$

ε: fractional excess density with respect to a reference density

b: apparent weight of a water particle of density ϱ surrounded by a medium having density ϱ_0

- Baroclinic

state of a fluid in which isopycnals and isobars are mutually inclined

An horizontal buoyancy gradient (baroclinic density field) contributes to the horizontal pressure gradient:

$$\frac{1}{\rho_0}\nabla_H p = \nabla_H \int_z^{\eta} \frac{g\rho}{\rho_0} dz = \underbrace{g\nabla_H \eta}_{\text{Barotropic}} - \underbrace{\int_z^0 \nabla_H b\, dz}_{\text{Baroclinic}}$$

Some definitions

Buoyancy Flux B.

Buoyancy is input to the ocean at the surface or laterally trough the heat or the fresh water flux.

$$B = \frac{g\alpha_T}{\rho c_p}\left(Q_s - Q_b - Q_h - Q_e\right) + g\beta\left(P - E - R\right)S$$

Where

$\left(Q_s - Q_b - Q_h - Q_e\right) = Q$ surface heat flux (see later for explanations)

P=precipitation

E=Evaporation

R=river runoff

α_T = Thermal expansion coefficient

β = Saline contraction coefficient

Some definitions

Brunt-Vaisala Frequency.

Water parcels displaced vertically are subject to a buoyancy restoring force that cause them to oscillate around the neutral stability level with a Frequency N:

$$N^2 = -\frac{g}{\rho_0}\frac{\partial \rho}{\partial z} = \frac{\partial b}{\partial z}$$

Richardson number

Vertical mixing can occur in presence of a vertical buoyancy gradient only if sufficient Kinetic energy is produced to overcome the buoyancy restoring force. A measure of this requirement is the Richardson number:

$$Ri = \frac{\partial b / \partial z}{(\partial u / \partial z)^2}$$

Turbulent mixing in general occurs for $Ri < 0.25$

Heat budget of oceanic and coastal areas

The temperature distribution of any region of the ocean is controlled by:

- Net Heat flux trough sea surface
- Heat exchange by advection and diffusion with adjacent regions.

Heat flux gain: solar radiation (received directly or reflected/scattered from clouds/atmosphere)
Heat flux losses longwave back radiation, latent and sensible heat flux

Advective changes (positive and negative) of heat content determined by currents (horizontal and vertical)

Diffusive heat flux due to horizontal/vertical temperature gradients.

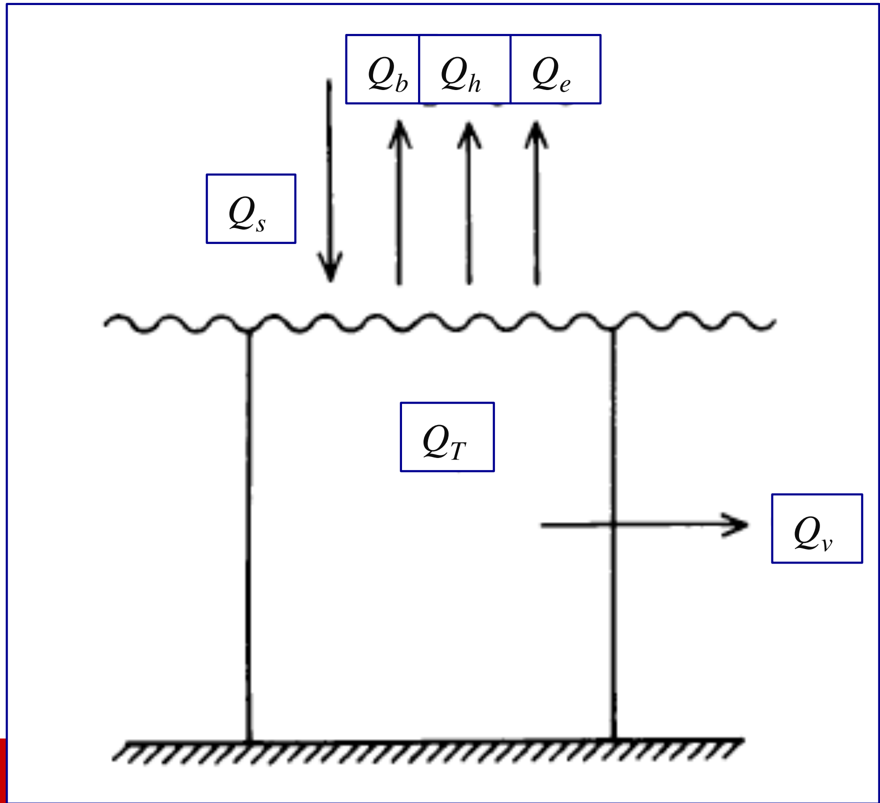
Vertical diffusive flux from small scale turbulent mixing or, if surface layer cooled, convective overturning

Horizontal diffusive heat exchanges from eddy motions at various space scales

Heat budget of oceanic and coastal areas

$Q_T = Q_s - Q_b - Q_h - Q_e - Q_v$ let also: $Q = Q_s - Q_b - Q_h - Q_e$

Q_s: Heat flux from incoming solar radiation.
Q_b: Heat flux lost by longwave (back) radiation
Q_h : Sensible heat flux (conduction)
Q_e: Latent heat flux associated to evaporation



Temperature response to Heat budget variation

Open ocean response.

Let δq be the quantity of heat absorbed by a unit volume of sea water

The corresponding temperature increase δT is given by: $\delta T = \dfrac{\delta q}{\rho c}$

c: specific heat (constant pressure)

If the heat gain occurs in a time δt , then the temporal rate of temperature increase is

$$\frac{\partial T}{\partial t} = \frac{1}{\rho c}\frac{\partial q}{\partial t}$$

The equation:

$Q_T = Q_s - Q_b - Q_h - Q_e - Q_v$

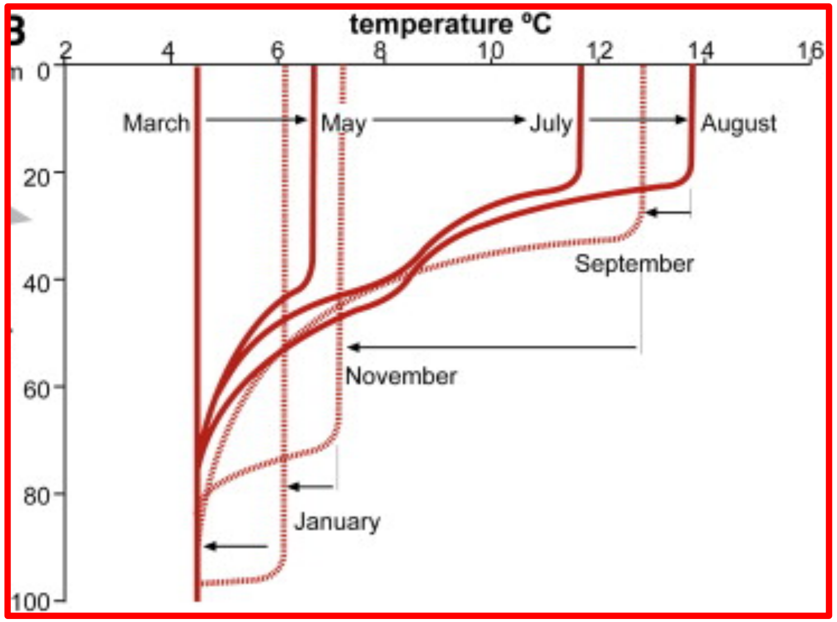
Denote the rate of gain of heat of a vertical seawater column (unit surface area).
If its height is h, then:

$$\frac{\partial q}{\partial t} = \frac{Q_T}{h} \quad \text{and} \quad \frac{\partial T}{\partial t} = \frac{1}{\rho c}\frac{Q_T}{h}$$

Temperature response to Heat budget variation

Open ocean response.

However, a positive net heat gain trough the surface (Q) is absorbed by a surface layer with a thickness of only a few metres, therefore for a positive Q only a thin surface layer become warmer and less dense than the water below.
Wave motion and wind stress generate turbulent energy to mix the heat input downward trough a well mixed layer to a depth extending several tens of metres.
Below such depth turbulence cannot mix the heat further down and a (seasonal) thermocline is formed at the base of the mixed layer.

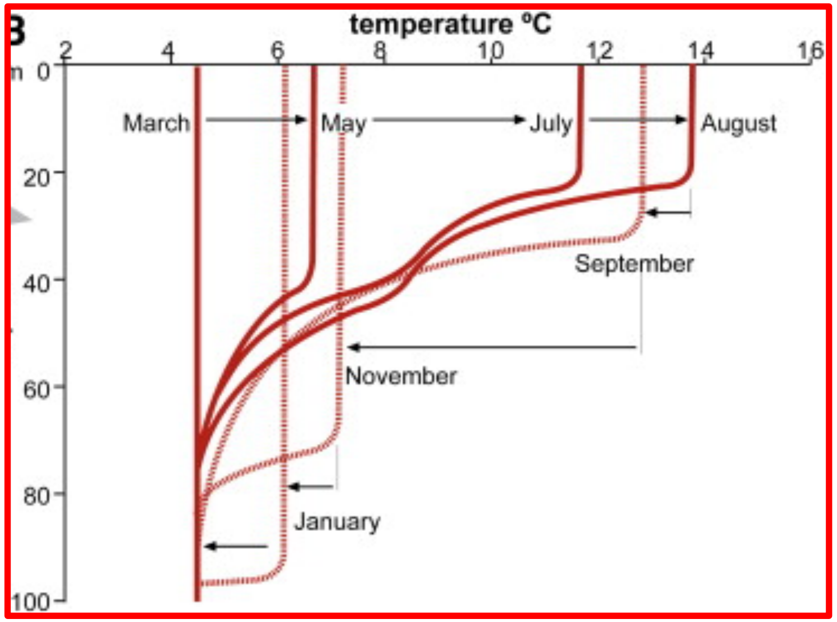


Temperature response to Heat budget variation

Open ocean response.

At temperate and polar latitudes the seasonal variation of Q determine alternating periods of positive and negative heat flux, involving the formation (summer) and destruction of the seasonal thermocline.
At tropical latitudes Q is permanently positive and therefore the "seasonal" thermocline is a permanent feature

Almost universal Occurrence in the mid-latitude open ocean

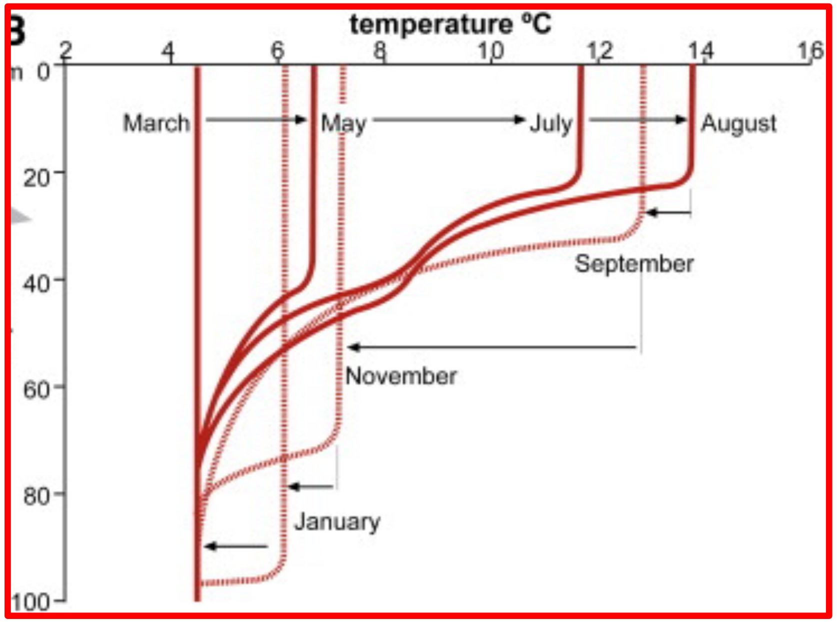


Temperature response to Heat budget variation

Coastal ocean response.

The process leading to the formation/destruction of the seasonal thermocline are acting also on the coastal ocean, but turbulence generated by the bottom friction (particularly in areas affected by significant tidal dynamics) may provide a sufficient level of energy to keep the water column always in well mixed conditions, and the development of the thermocline depend mostly on the bottom depth and on the levels of mixing.

it has to be considered that shallower water column are heated/cooled up fastly than deeper and the interplay between mixing and heating can generate very different conditions.

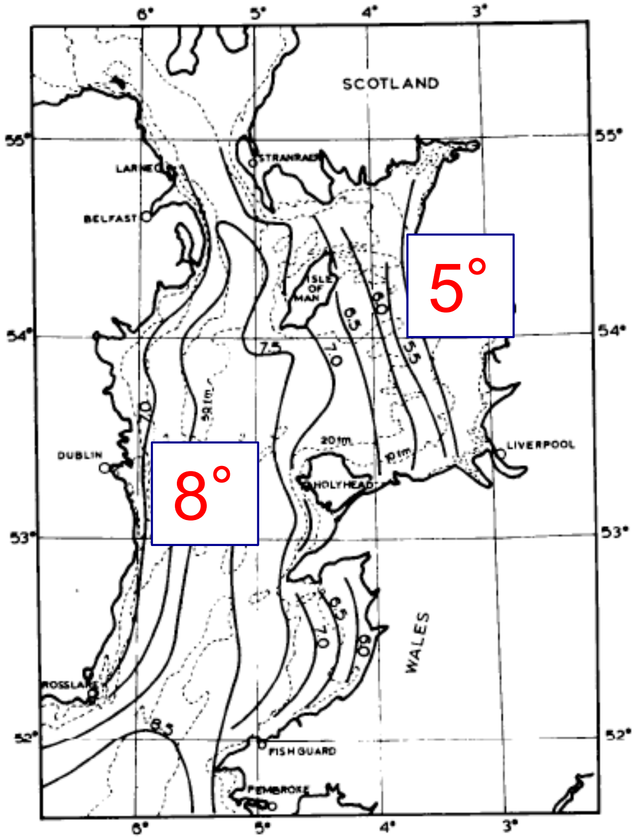


Temperature response to Heat budget variations

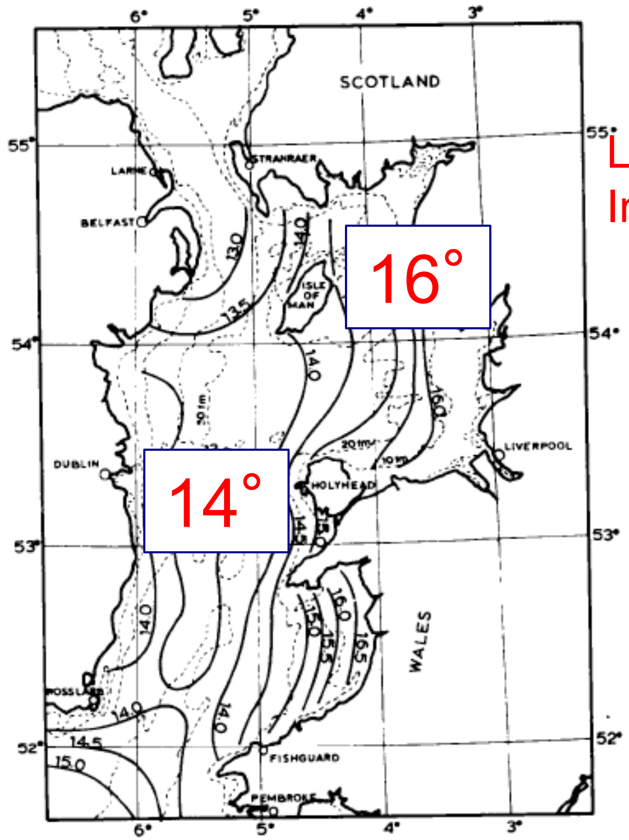
Irish sea

larger cooling
In coastal areas

WINTER



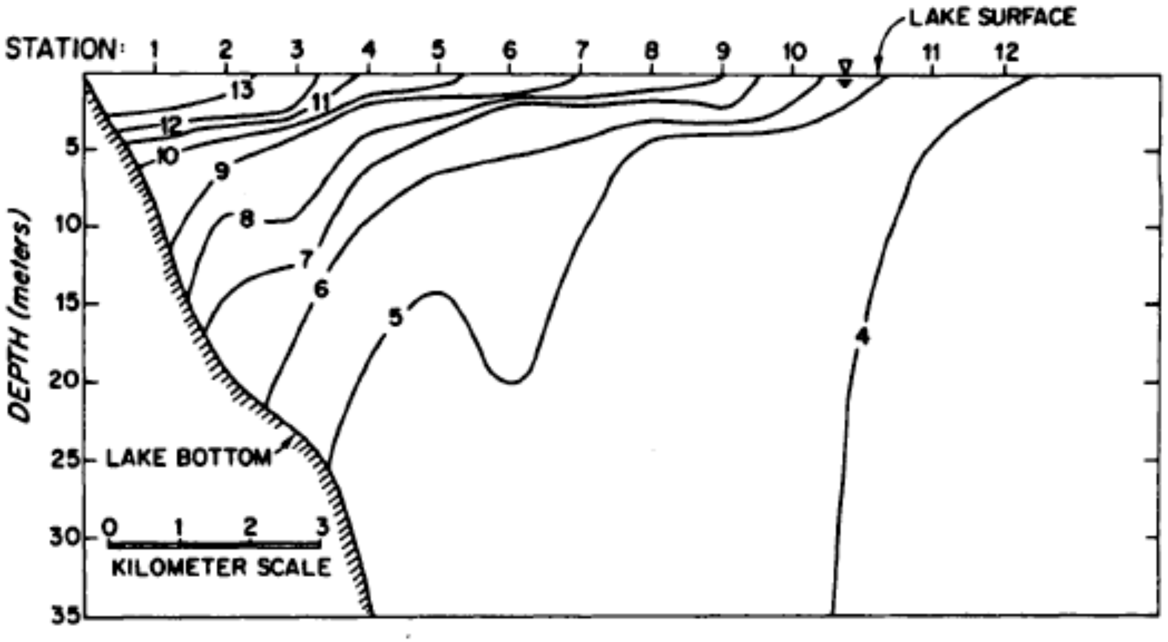
SUMMER



Larger warming
In coastal areas

Stratification in the coastal ocean

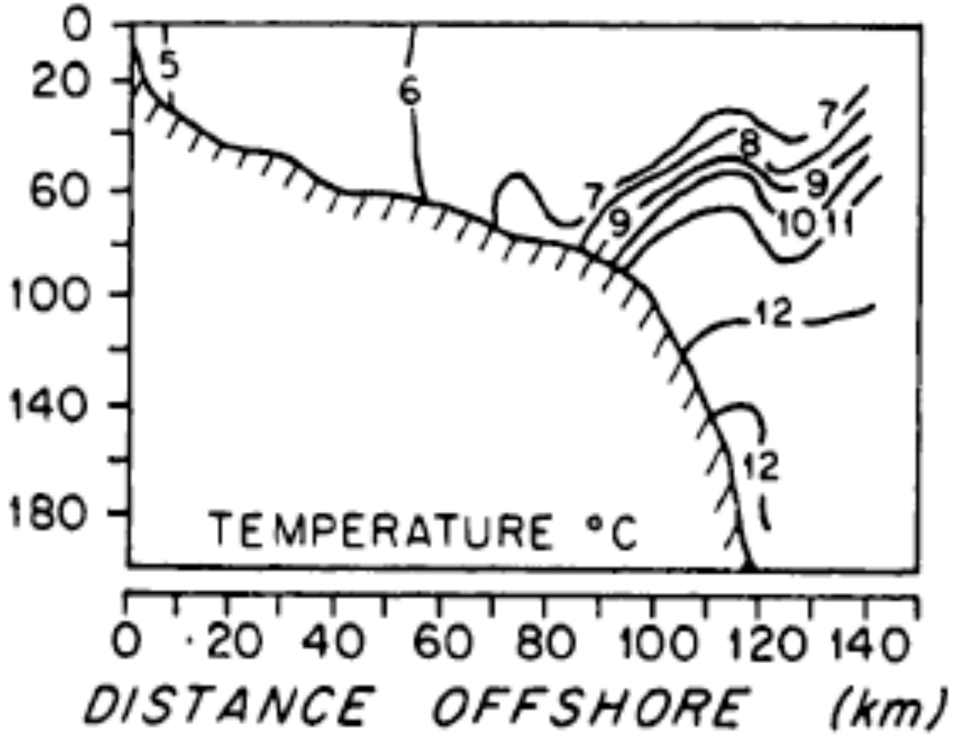
Areas with reduced mixing:
Stronger thermocline in coastal areas with respect to the off-shore



Temperature section

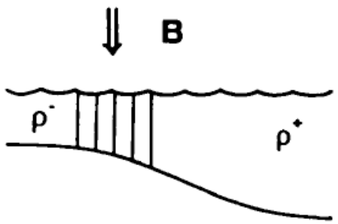
Stratification in the coastal ocean

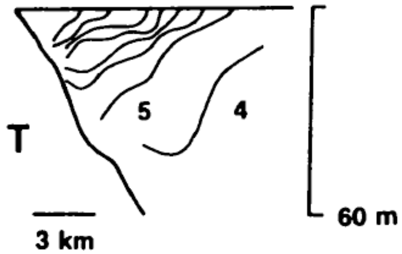
Areas with strong mixing.
Coastal zone (under cooling processes) is kept well mixed while the offshore develop stratification



Stratification in the coastal ocean

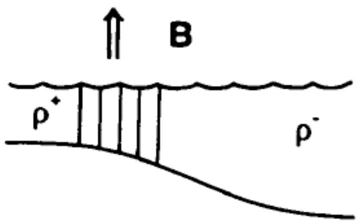
$\Downarrow B$=surface Buoyancy flux (input) trough heat (warming) or freshwater (precipitation)

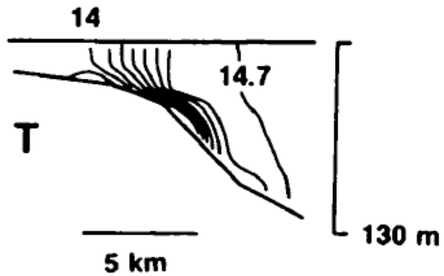




Buoyancy input prevail over mixing
Stratification develops

$\Uparrow B$=surface Buoyancy flux (extraction) trough heat (cooling) or freshwater (precipitation)





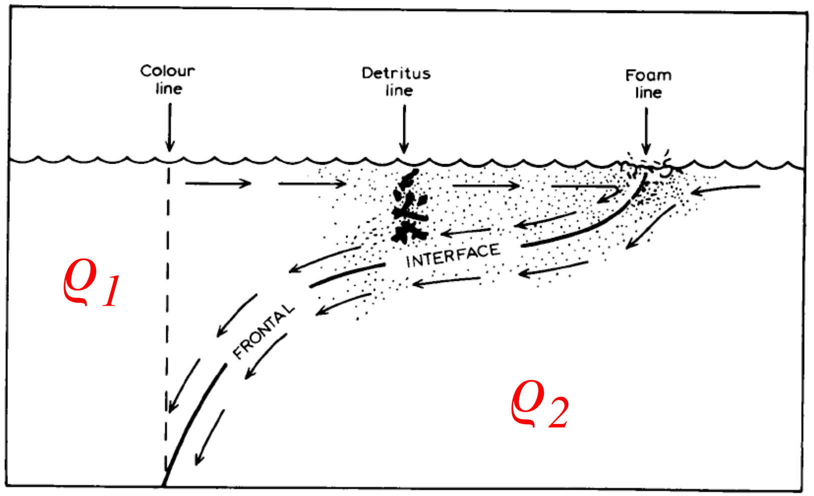
Buoyancy extraction prevail. Dense water are formed.

Thermohaline circulation and fronts

Horizontal patterns of temperature distributions (as well as salinity's…seen later) originates Density (and hence pressure) horizontal gradients, giving rise to a dynamic that, since it depends on heat and fresh water fluxes is defined as "Thermohaline",

The differential heating/cooling of the coastal water with respect to the offshore generate a well defined structure called "front", mostly characterised by a sharp horizontal density difference, Determined by temperature and/or salinity, but also marked by other characteristics such as:

- Change of seawater color due to phytoplankton and detritus accumulation on one side of the front
- Foam accumulation near the front

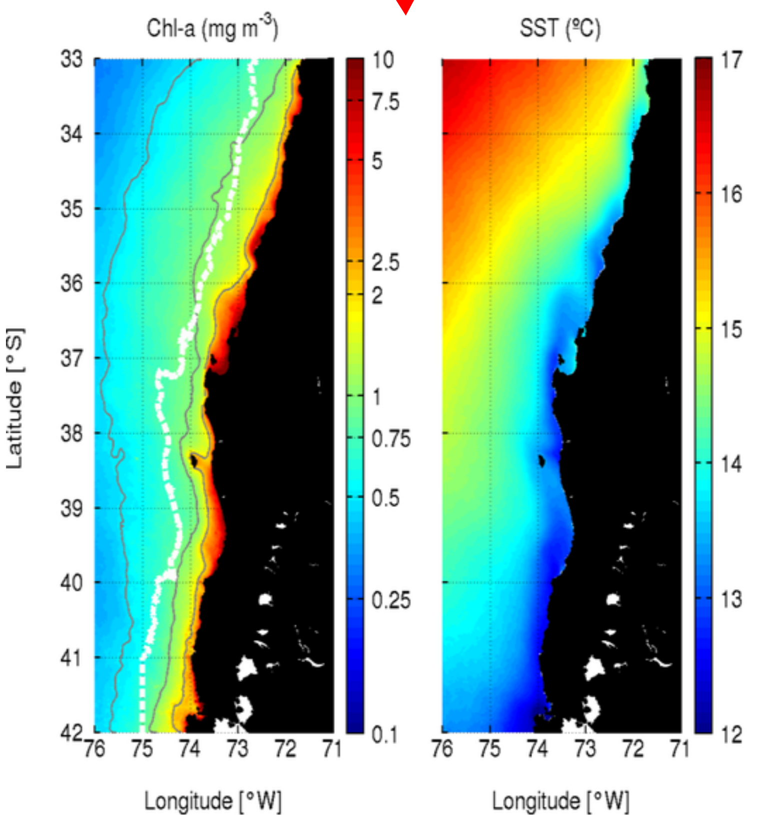


$\varrho_1 < \varrho_2$

Fronts

Temperature and Phytoplankton gradients

Color changes marked by A foam line

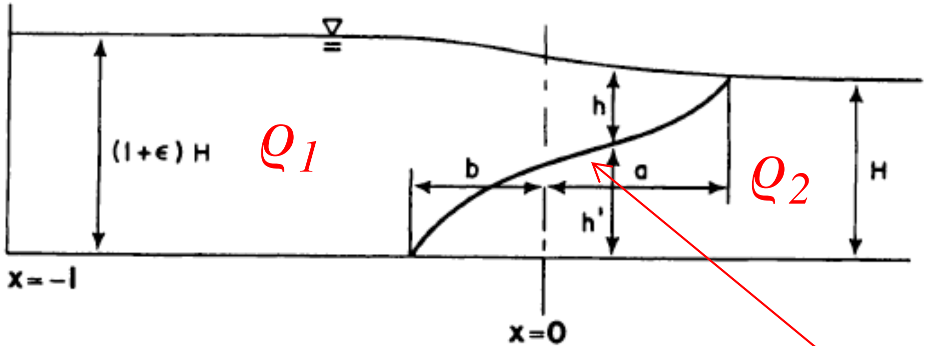


Fronts

Front features are determined by the currents patterns in its vicinity

Consider the vertical structure shown below (having depth H)

Shore

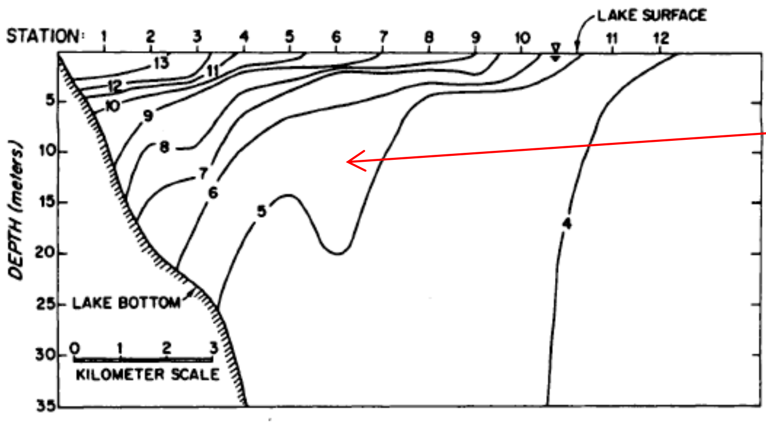


Offshore

$\varrho_1 < \varrho_2$

Resembling in an idealised way the structure previously seen.

Shore

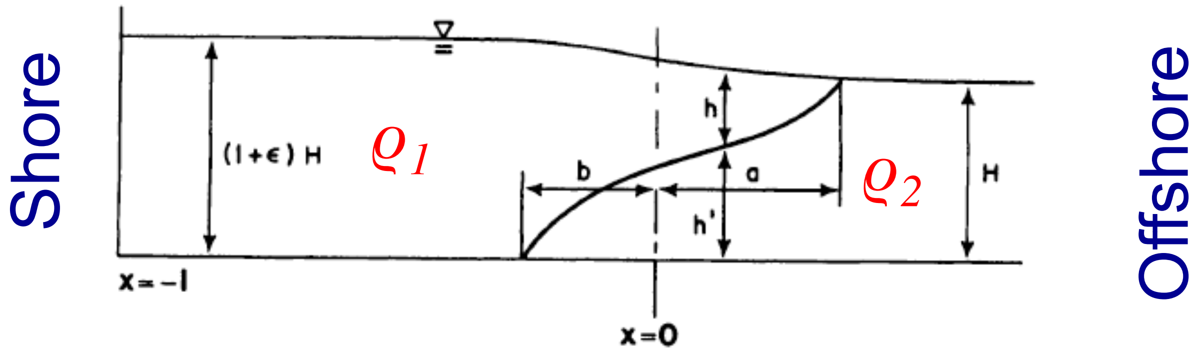


Offshore

An idealised
S-shaped interface
Marking the contact
Between water with
Different density

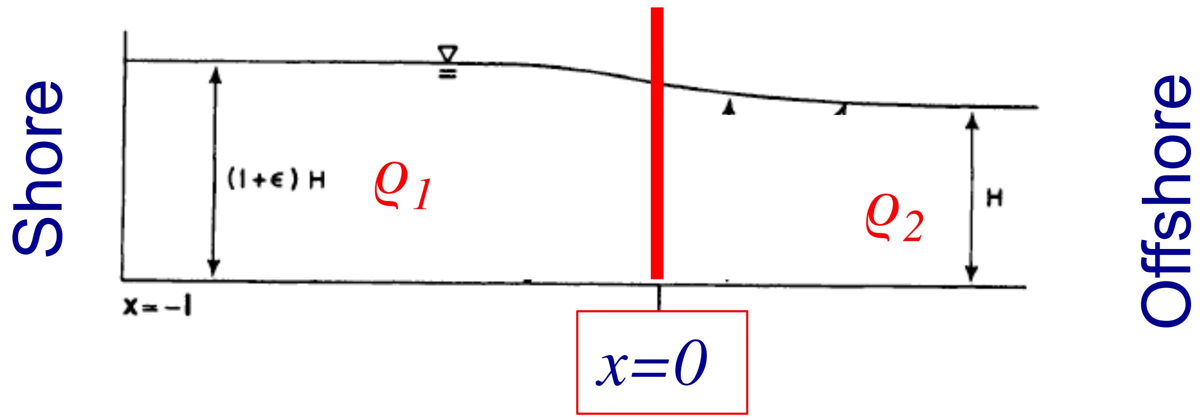
Fronts

This structure

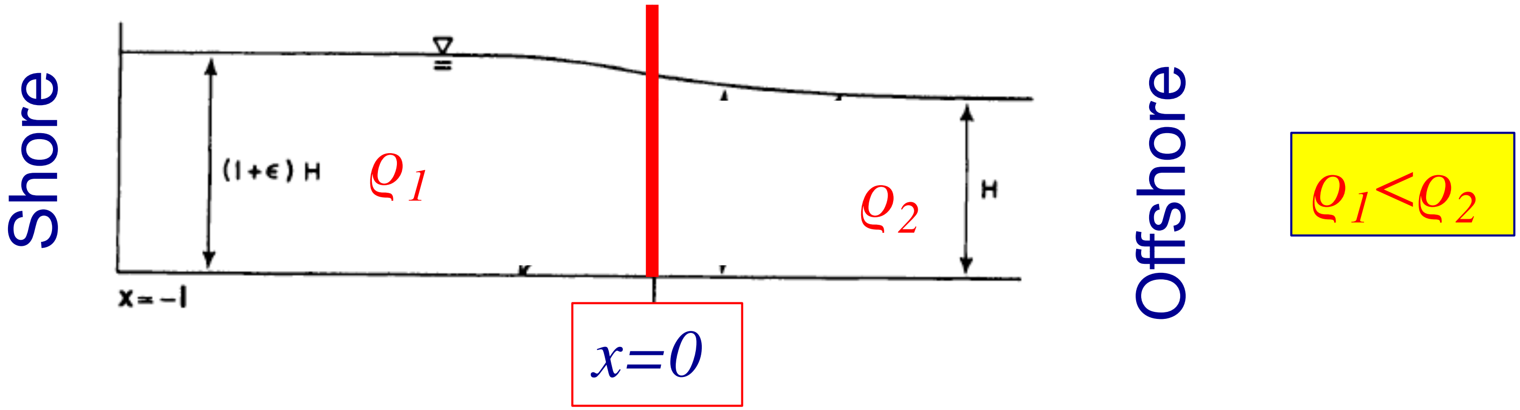


$\varrho_1 < \varrho_2$

Originating from an initial situation determined by a membrane inserted into the fluid at $x=0$.



Fronts



Density in the shore side region ($-l \leq x \leq 0$) is lower than the density in the offshore ($x \geq 0$) region due to stronger heating and/or dilution processe

The fractional density difference ε is given by: $\varepsilon = \dfrac{\rho_2 - \rho_1}{\rho_2}$

The fluid depth in the "shore" region therefore increase from H to $H(1+\varepsilon)$.

The lighter (heated and/or diluted) stands higher on one side of the membrane and therefore has an excess potential energy per unit mass.

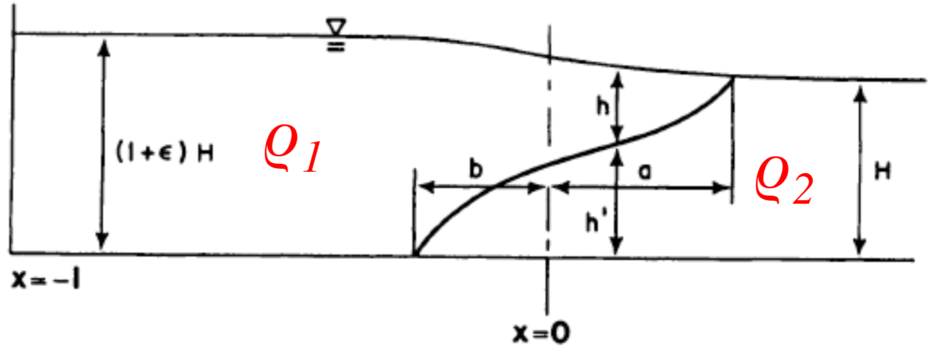
$$\Delta E_p = \frac{1}{2} g \varepsilon H = \frac{1}{2} g' H$$

Fronts

Assume that (at $t=0$) the "membrane" is removed allowing water to move freely (but without mixing and friction) between the "heavy" and "light" columns.

$\varrho_1 < \varrho_2$

Shore



Offshore

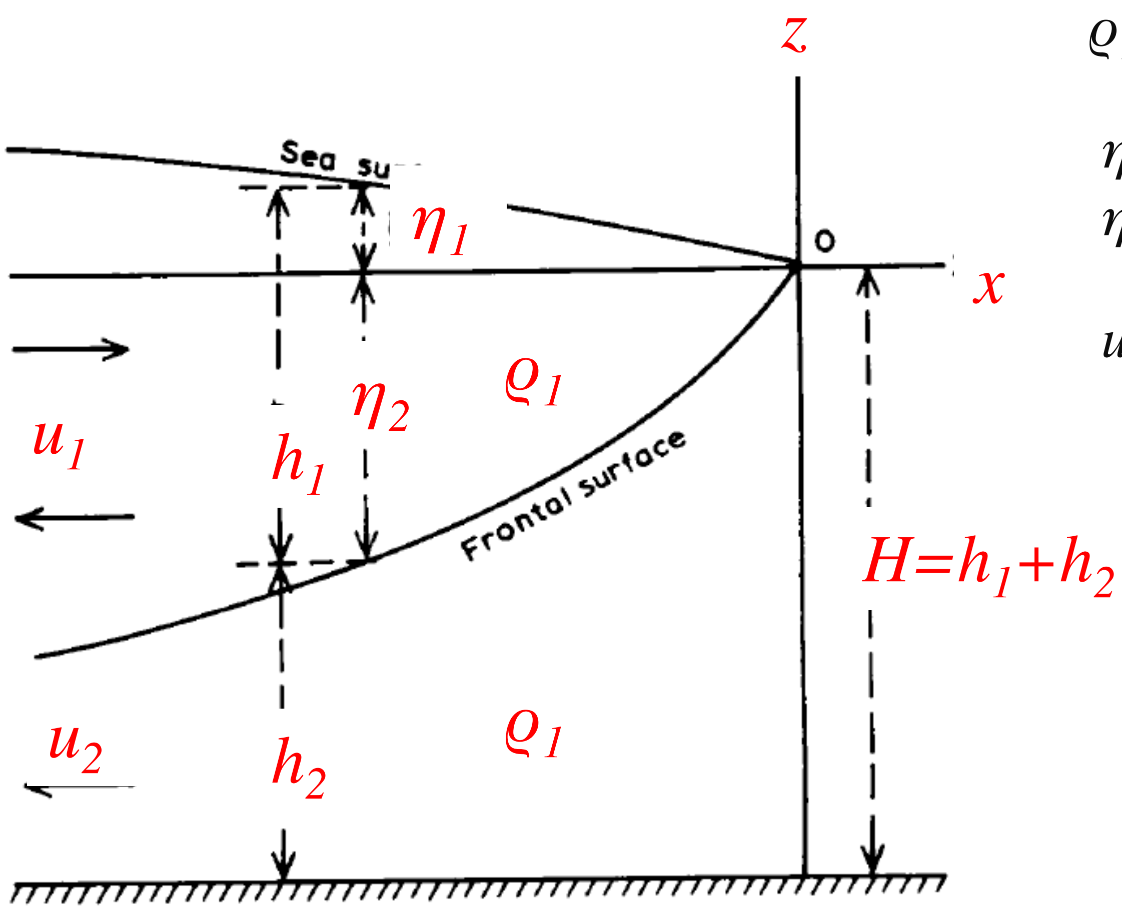
Upon membrane removal motion starts with lighter water moving at "surface" upon denser water and with denser water moving at "depth" below the lighter water, thereby originating the S-shaped interface among different density water.

The motion velocity c can be computed assuming conversion from Potential to kinetic Energy:

$$c = \left(\varepsilon g H\right)^{1/2} = \left(g' H\right)^{1/2}$$

Fronts

Let references axes be taken as in the figure below (origin at the front , x-axis perpendicular to the front, y-axis along it and z-axis vertically pointing upward.



ϱ_1, ϱ_2: density in the upper and lower layer (respectively.)

η_1: free surface elevation

η_2 : elevation (≤0) so that $\eta_1 - \eta_2 = h_1$, the thickness of the upper layer

u_1, u_2: cross frontal velocities in the upper and lower layer (respectively)

Coriolis and advective terms are disregarded….
Eddy viscosity is assumed constant
And exchanges between layers
Is assumed non-existent.

Fronts

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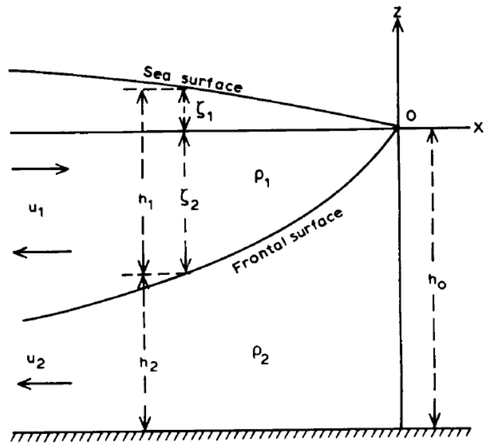

Under such assumptions, the equation of motion in the cross frontal direction reduces to:

$$0 = -\frac{1}{\rho_1}\frac{\partial p_1}{\partial x} + K_z \frac{\partial^2 u_1}{dz^2} \qquad \frac{\partial p_1}{\partial x} = \rho_1 K_z \frac{\partial^2 u_1}{dz^2}$$

$$0 = -\frac{1}{\rho_2}\frac{\partial p_2}{\partial x} + K_z \frac{\partial^2 u_2}{dz^2} \qquad \frac{\partial p_2}{\partial x} = \rho_2 K_z \frac{\partial^2 u_2}{dz^2}$$

Where $K_z = A_v/\varrho_0$

Fronts

Coriolis and advective terms are disregarded….
Eddy viscosity is assumed constant
And exchanges between layers
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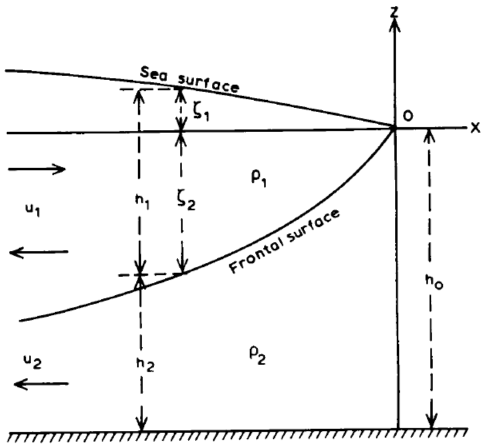
$$\frac{\partial p_1}{\partial x} = \rho_1 K_z \frac{\partial^2 u_1}{dz^2}$$

$$\frac{\partial p_2}{\partial x} = \rho_2 K_z \frac{\partial^2 u_2}{dz^2}$$



Where pressures p_1 and p_2 are given by the hydrostatic equation:

$$\frac{\partial p_1}{\partial x} = g\rho_1\left[\eta_1 - z\right]$$

$$\frac{\partial p_2}{\partial x} = g\rho_1\left[\eta_1 - \eta_2\right] + g\rho_2\left[\eta_2 - z\right]$$

Fronts

Coriolis and advective terms are disregarded….
Eddy viscosity is assumed constant
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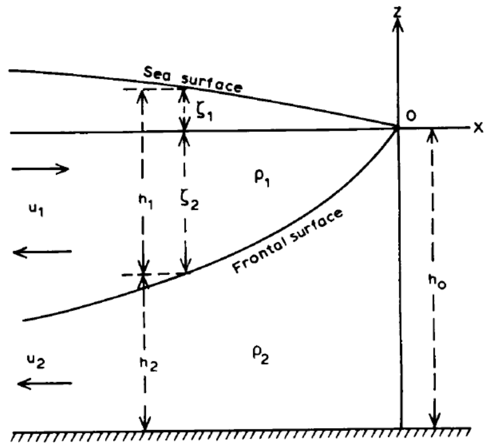
$$\frac{\partial p_1}{\partial x} = \rho_1 K_z \frac{\partial^2 u_1}{dz^2}$$

$$\frac{\partial p_2}{\partial x} = \rho_2 K_z \frac{\partial^2 u_2}{dz^2}$$



Continuity equation for the two layers yields

$$\int_{-h_1}^{0} u_1 dz = 0$$

$$\int_{-H}^{-h_1} u_2\, dz = q = -cH$$

constant

N.B: The further assumption

$$\eta_1 << \eta_2$$

Is made, so that the limit of integration η_1 is taken as 0, and $\eta_1$as h_1.

no net integrated flow in the upper layer, constant flow in the lower layer.
if the front is advancing with velocity c (defined previously) over water initially at rest then:

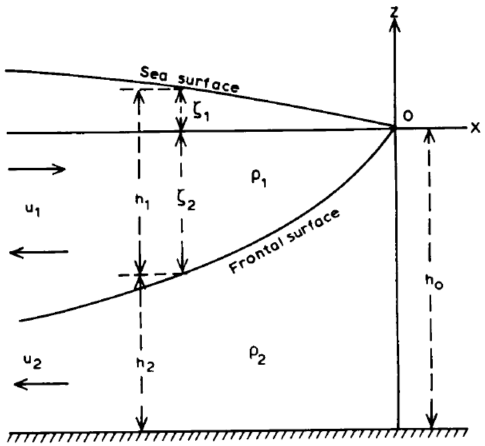
$$q = -cH$$

Fronts

Coriolis and advective terms are disregarded….
Eddy viscosity is assumed constant
And exchanges between layers
Is assumed non-existent.

$$\frac{\partial p_1}{\partial x} = \rho_1 K_z \frac{\partial^2 u_1}{dz^2}$$

$$\frac{\partial p_2}{\partial x} = \rho_2 K_z \frac{\partial^2 u_2}{dz^2}$$



The further assumption are:

$h_1 << H$ $\quad \partial u_2 / \partial z = 0$

Velocity in the lower layer vertically uniform.

The stress at the interface is related to velocity in the Lower layer trough a quadratic law with a drag coefficient c_d.

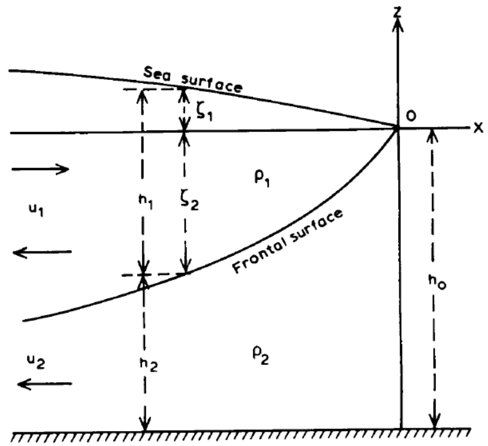
Therefore the boundary conditions are as follows:

Fronts

Coriolis and advective terms are disregarded….
Eddy viscosity is assumed constant
And exchanges between layers
Is assumed non-existent.

$$\frac{\partial p_1}{\partial x} = \rho_1 K_z \frac{\partial^2 u_1}{dz^2}$$

$$\frac{\partial p_2}{\partial x} = \rho_2 K_z \frac{\partial^2 u_2}{dz^2}$$



Therefore the boundary conditions are as follows:

$$\left.\frac{\partial u_1}{\partial z}\right|_{z=0} = 0$$

$$u\big|_{z=-h_1} = u_2 \qquad \text{or also} \qquad A_v \left.\frac{\partial u_1}{\partial z}\right|_{z=-h_1} = c_d u_2^2$$

Fronts

Coriolis and advective terms are disregarded….
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And exchanges between layers
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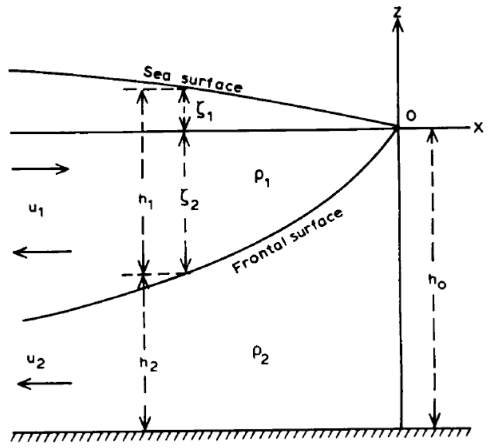
$$\frac{\partial p_1}{\partial x} = \rho_1 K_z \frac{\partial^2 u_1}{dz^2}$$

$$\frac{\partial p_2}{\partial x} = \rho_2 K_z \frac{\partial^2 u_2}{dz^2}$$



Solution for the system are:

$$u_1 = u_s\left(1 - 3\frac{z^2}{h_1^2}\right)$$

with $u_s = \dfrac{3\rho_1 K_z}{2\rho_2 c_d h_1}$

$$u_2 = -2u_s$$

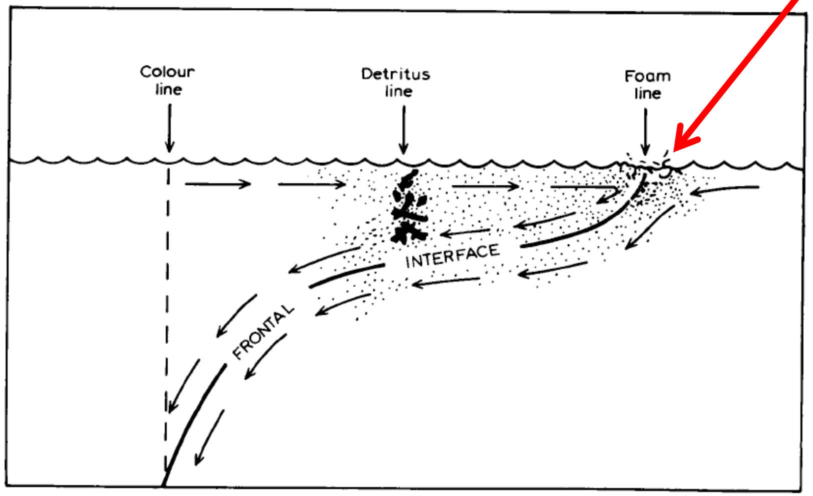
velocity at the interface is opposite to the surface velocity ($z=0$) and twice in magnitude.

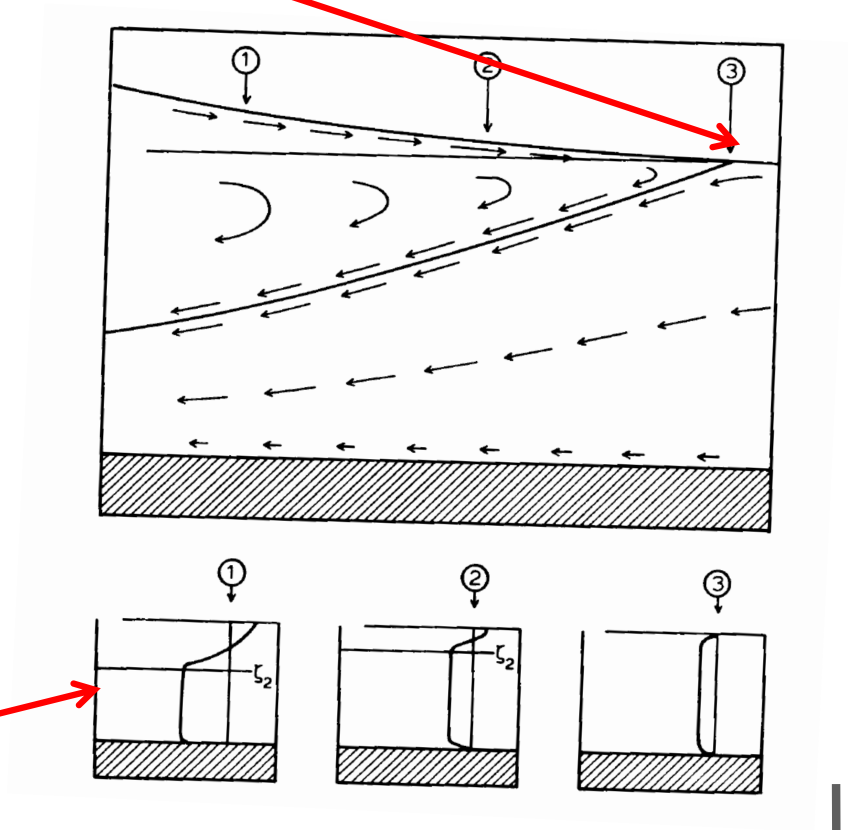
The form of the interface $\eta_2(x)$ is described by

$$x = \frac{c_d \varepsilon \eta_2^4}{36 K_z}$$

Fronts

Convergence (accumulation of detritus)





Velocity profiles at various distances from the front